S T U D I A G E O M O R P H O L O G I C A C A R P A T H O - B A L C A N I C A
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L A N D F O R M E V O L U T I O N I N M O U N T A I N A R E A S
Recent geomorphological hazards in Carpatho-Balcan-Dinaric region

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THRESHOLD VALUES OF RAINFALLS TRIGGERING SELECTED DEEP-SEATED LANDSLIDES IN THE POLISH FLYSCH CARPATHIANS

Abstract. This paper presents the influence of rainfalls, their sums, intensity and duration, on mass movement processes exemplified by selected landslides in the Flysch Carpathians. The effort is made to find "rainfall thresholds" which, when exceeded, cause mass movements to develop rapidly. Determination of the thresholds for various environmental conditions of the Carpathians is attempted. The mutual influence of rainfall and water circulation in the ground (substratum) on conditions initiating the mass movements has been emphasised.

Key words: rainfall thresholds, landslide, Flysch Carpathians

INTRODUCTION

The morphogenetic role of precipitation and the resulting water circulation are reflected in relief modelling in various ways and with various intensities. The morphogenetic role of rainfalls depends on their sums as well as on their temporal and spatial distribution. In general, it depends on many factors, i.a. — geological conditions, relief and its morphometric features, land use etc.

We can talk about relief-forming processes, especially about the extreme ones controlled by rainfalls, when the critical values of rainfalls — sums, intensities, duration — are exceeded. These relief-forming processes should be considered at the background of microclimatic zones and domains, in which various geoecosystems have developed and which show a different sensitivity to changes (Starkel 1976, 1996).

In the Flysch Carpathians the landsliding plays a crucial role in modelling the relief and a whole natural environment. On the other hand, the landsliding is destructive and difficult for land management. In the Polish Carpathians, ca 20,000 landslides have been inventoried which amounts to 20% of this region area (*Katalog osuwisk*… 1975; Mrozek et al. 2000).

Meteorological (rainfall) and hydrological (water circulation in substratum) conditions (Kleczkowski 1955; Thiel 1976; Gil and Kotarba 1977; Thiel 1989; Gil 1994; Rączkowski and Mrozek 2002; Zabuski et al. 2004) are exceptionally important in slope modelling by mass movements in the Flysch Carpathians. The relationship between landsliding and rainfall has been a subject of numerous studies.

N. Caine (1980), based on 73 hydrometeorological observations collected in areas with diversified relief, geological conditions and located in various climatic regions developed an equation relating the average rainfall intensity (I) and its duration (D). The threshold values were expressed by the formula: $I = 14.82 \cdot D^{-0.39}$. After exceeding the thresholds, debris flows and shallow landslides (to 3 m deep) could have been formed. J. L. Innes (1983) in his equation considered the rainfall sum (instead of rainfall intensity) and rainfall duration. The critical value was described as: $T = 4.9355 \cdot D^{0.5041}$, where T was the rainfall sum (in mm) and D was rainfall duration (in hours).

M. Govi et al. (1982), based on the studies performed in Piedmont, stated that shallow mass movements were set off by 1–3 day critical rainfall, which made 13–14% of the annual precipitation totals. The critical precipitation depends on precipitation which occurred in a 30–40 day long preceding period. M. Govi et al. (1982) called this period “dry” if the rainfall amounted to 70 mm and “wet” if the rainfall amounted to 140–300 mm. A common occurrence of mass movements was observed, when critical rainfall amounted 28 to 30% of the annual precipitation total and rainfall intensity in the final 3–6 hour long phase was 30–40 mm $\cdot$ h^{-1}.

R. Giannecchini (2005) stated, that shallow landslides in the Apennines were triggered by very intensive rainfalls of the order of 325 mm $\cdot$ 4 h^{-1} with the maximum intensity of 158 mm $\cdot$ h^{-1}. The shallow landslides also developed under the influence of less intensive but longer-lasting precipitation — of the order of 160 mm $\cdot$ 13 h^{-1} with intensity of 30 mm $\cdot$ h^{-1}. Continuous rainfalls with monthly totals exceeding 600 mm drastically decreased the critical values which triggered such type of landslide.

M. T. J. Terlin (1997), based on his observations in the Columbian Andes (Manizales region) in 1993, showed that shallow, weathered material landslides were triggered by diurnal precipitation of 70 mm while the deep-seated ones were initiated during precipitation amounting to 200 mm in 25 days preceding the movement and when daily precipitation values were 0–50 mm.

The studies performed in New Zealand (Glade 1998) allowed determining two precipitation thresholds initiating shallow landslides and debris flows. The first (minimum probability) threshold, which denoted the daily precipitation total below which the landsliding processes in Wellington were not observed, was 20 mm. The second (maximum probability) threshold occurred when precipitation was above 140 mm, and then landsliding was always observed.

W. Froehlich and L. Starkel (1987, 1991), studying these processes in the Darjeeling Himalaya, stated that the diurnal precipitation exceeding 250 mm and a few day long precipitation exceeding 350 mm triggered local debris flows while the common transformation of slopes took place when 2–3 day long precipitation exceeded 600 mm with the rainfall intensity reaching to 50 mm $\cdot$ h^{-1}.

Based on the studies carried out in the Flysch Carpathians it became evident that precipitation totals belong to the major causes of mass movements. K. Jakubowski (1964, 1965), E. Gil and L. Starkel (1979) accepted that precipitation exceeding 100 mm is critical. K. Thiel (1976) used both precipitation total and precipitation frequency to determine indices of "precipitation concentration". The indices were defined as quotients of precipitation total in a given period and an appropriate mean precipitation total of a multi-annual period. Following this way of reasoning, the annual precipitation concentration index is defined as the quotient of annual precipitation total in a given year and the mean annual precipitation total of many years (10–20 years or more). Appropriately, the monthly precipitation concentration index is the precipitation total of the given month divided by the mean monthly precipitation of the multi-annual period. The daily precipitation concentration index is analogous and refers to daily precipitation totals.

From the detail studies carried out at "Bystrzyca" landslide (Gil and Starkel 1979; Thiel 1989) it became evident that the landslides in the Carpathians formed when the annual precipitation totals exceeded 1,000 mm, monthly totals exceeded 200 mm, when precipitation with intensity below 0.025 mm $\cdot$ min^{-1} prevailed and during long-lasting rainy periods summing up to 260–280 mm. These last rainfall values correspond well with a complete saturation condition, which is already satisfied if rainfall exceeds 160 mm.

Similar relations presented E. Gil (1994, 1997) who analysed some Carpathian landslides and concluded that in the regions where shales prevailed, landslides developed after 20–45 day long rainy periods during which rainfalls with the average intensity of 0.025 mm $\cdot$ min^{-1} gave the sum of 200–300 mm. In the regions where sandstones prevailed the precipitation totals amounted to 400–500 mm, but in the last 5–6 days of the rainy period the total was ca 250 mm. Moreover, if the totals were ca. 200–300 mm in the rainy periods, the movements were rejuvenated on periodically active landslides.

L. Starkel (1996) related three types of mass movements to three types of rainfalls:

1. Short-lasting downpours amounting to 30–70 mm, whose intensities were of 1–3 mm $\cdot$ min^{-1}. Such downpours, of local extents, resulted in overland flow, slope-wash, mudflows, gully erosion and accumulation on valley floors.
2. Continuous rainfalls which lasted 1 to 5 days, with an average intensity not exceeding 3–25 mm $\cdot$ h^{-1} and amounting to 150–400 mm. Such rainfalls led to soil saturation with water and to rise in groundwater level. As the rainfall continued the surplus water flowed out, and the landslides and extreme floods were formed.

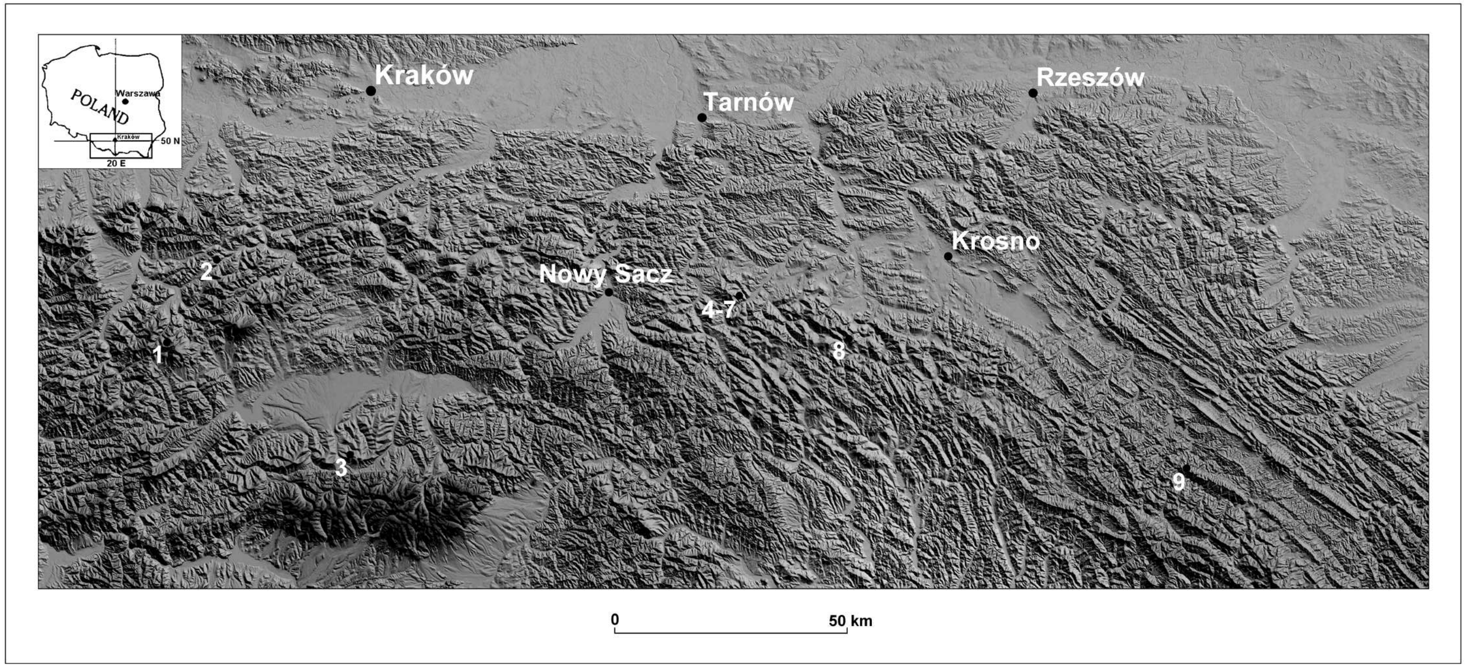

Fig. 1. Location of selected landslides at the background of the relief of the Carpathians. Fragment of DTM of Poland (Silesian University). Landslides: 1 — landslides on Pilsko (Beskid Żywiecki Mts.), 2 — Lachowice (Beskid Żywiecki Mts.), 3 — Kotelnica (Gubałówka Foothills), 4–7 — Bystrzyca, Szymbark Folwark, Szymbark Szklarka (Maślana Góra), Bystra Polesie (Beskid Niski Mts.), 8 — Lipowica (Beskid Niski Mts.), 9 — Połoma (Bieszczady Mts.)

3. Rainy seasons, when annual precipitation totals reached 1,500 mm, provided the rainfall of summer season amounted to 200–500 mm. Slope covers became water saturated. Rainfall intensity was low and did not exceed 1–1.5 mm $\cdot$ h^{-1}. Under such circumstances, increase in rainfall intensity triggered landsliding. Deep-seated, rock landslides were formed.

E. Gorczyca (2004), in order to define the threshold values, introduced a term — "an activating impulse". Based on her studies in the Beskid Niski Mts., she decided that 2 hour long rainfall amounting to 70–150 mm that followed the 6 day long continuous rainfall with the average intensity of 0.123 mm $\cdot$ min^{-1} was decisive of landsliding in 1997. The most numerous were shallow movements which was a characteristic feature of downpours.

W. Rączkowski and T. Mrozek (2002), when analysing the mass movements in the Carpathians in the last 100 years, stated that formation of deep-seated rock landslides was preceded by long-lasting precipitation whose monthly totals reached 400–550 mm resulting in the full saturation of the ground.

The purpose of this paper is to show the role of precipitation in formation of new and reactivation of the old large, deep-seated landslides. In order to fulfil this task a number of landslides representing different regions of the Flysch Carpathians have been selected: Kotelnica (Gubałówka Foothills), Szymbark — Maślana Góra, Lipowica, Bystrzyca, Bystra Polesie, Szymbark Folwark (Beskid Niski Mts.), Połoma (Bieszczady Mts.) and landslides in Beskid Żywiecki Mts. (Fig. 1).

CHARACTERISTICS OF SELECTED DEEP-SEATED CARPATHIAN LANDSLIDES

In order to determine the influence of precipitation on the formation of new and rejuvenation of old landslides, the reliable data on precipitation from the nearest located gauging stations and on landslide movements are needed.

Despite of numerous landslides known in the Carpathians, the satisfactory information on the dates of their formation is difficult to obtain. The analysis carried out in this paper is based on the selected cases for which information about the mass movement processes is adequate (Fig. 1). The deep-seated structural landslides (Zabuski et al. 1999) which comprise earth-rock material are analysed. These landslides represent various geomorphologic regions and tectonic units of the Flysch Carpathians.

The landslide in the Szklarka valley near Szymbark, on the southern slopes of Maślana Góra (Fig. 2) was formed after continuous spring–summer rainfalls in 1913 (Sawicki 1917). The thorough examination and measurements performed by L. Sawicki were the first so detailed studies on the landslide processes in the Polish geomorphologic literature. The landslide was found in the marginal, tectonically deformed part of the Magura nappe (thrusts and scaling). The upper part of the landslide occurred on the southern flank of the Maślana Góra syncline which was composed of the Magura sandstones while its middle and lower parts,

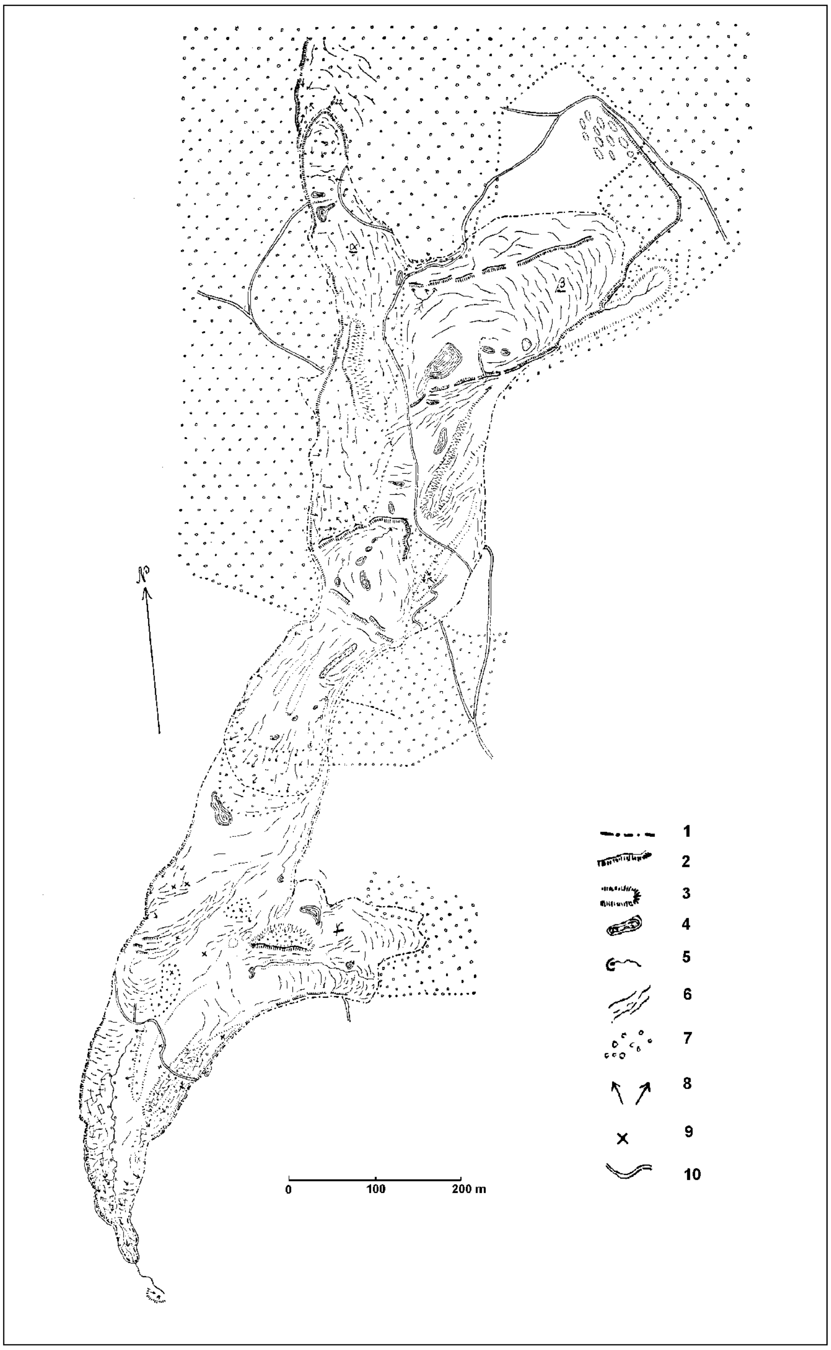

Fig. 2. Sketch of Szymbark Szklarka landslide on Maślana Góra (after Sawicki 1917). 1 — landslide boundary, 2 — edges of cracks and minor scarps, 3 — lateral and longitudinal bulges, 4 — ponds, 5 — springs, 6 — transverse cracks, 7 — forest, 8 — direction of fallen trees, 9 — damaged buildings, 10 — roads

dissected by a transverse thrust rested on shale-sandstone Inoceramian beds and Eocene variegated shales and Ciężkowice sandstones (S i k o r a 1970). The landslide was set off in a spring pot, affected the adjacent slopes and run out on to the Szklarka valley floor.

That was a complex structural landslide, comprising both weathered material and rocky substratum. In the head part the movement was rotational and then downward it transformed into translational one, thus the slide changed in an earth-debris flow.

L. S a w i c k i (1917) reported that the landslide was 1,970 m long, and its width ranged from 350 to 400 m at the head scarp, and to 250 to 350 m in the middle and toe parts. The height difference was 252 m (descend from 622 to 370 m a.s.l.). The movement of 1913 affected the whole body of the old landslide, widened it, and damaged the area of 42.8 ha. The maximum slip surface was at the depth of 20 m while the thickness of slid down material reached up to 15 m which gave an estimate of 3.4 million m^3 capacity. The landslide masses were transferred over the distance of 375 m. The landslide damaged 8 buildings.

The direct causes of landslide movements were high and long-lasting rainfalls which occurred in 1913 (Fig. 3).

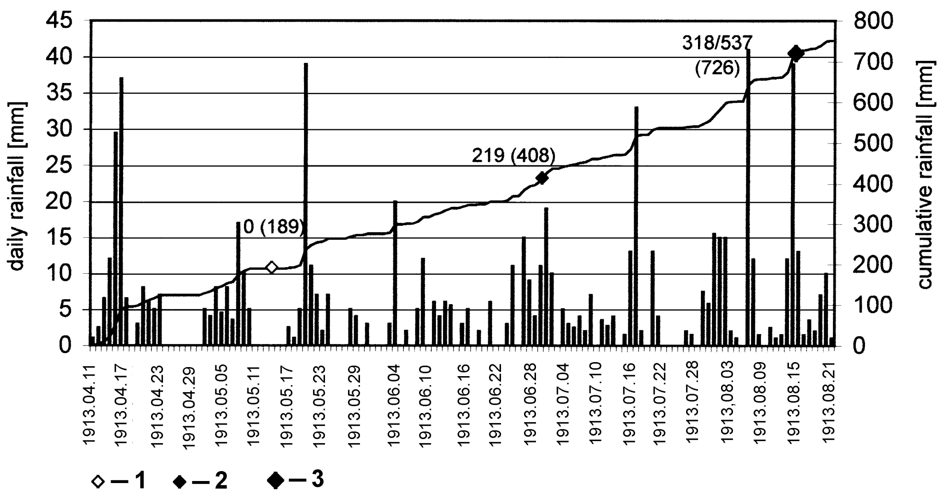

Fig. 3. Course of precipitation preceding the formation of Szymbark Szklarka landslide, 30.06.1913; 15.08.1913 — Grybów. 1 — onset of preceding rainfall, 2 — beginning of landsliding, 3 — main phase of movement

Rainfalls in spring–summer period of 1913 were exceptional. L. S t a r k e l (1996) calls such periods "rainfall seasons". Form the first 10-days of April until the end of the second week of August the precipitation total was 755 mm (Grybów). From 11 April until 10 May the precipitation reached 189 mm. This precipitation occurred as two ten-days series of continuous rainfalls or occasional showers (whose diurnal sums exceeded 10 mm) and, then, was followed by 6–8 day long intervals without rain. Since 16 May till the end of August, the frequency and height of daily precipitation increased, especially in the case the rainfalls exceeding 10 mm (the rains were paused for 1–2 days).

The landslide formed in two stages. The first movements were recorded by the end of June (Szymbark Parish Book), when the precipitation total from 16 May to 30 May was 219 mm (Fig. 3), but while calculated for the period from the beginning of a rain season, that is since 11 April, this precipitation sum was 408 mm. The latter was accepted as the precipitation initiating the landsliding. The movement comprised the upper part of the landslide body (head part), directly below the ridge of Maślana Góra. L. Sawicki (1917) described this material as thick loamy-clayey-shale material.

The main, catastrophic movements occurred in the first and second ten days of August 1913, when the precipitation total of the period form 1 July to 15 August was 318 mm, but counting form 15 May, being the "zero point" — 537 mm. Summing from the onset of the "rainy season", that is from 11 May to 15 August, the precipitation total was 726 mm.

The landslide front stabilised at the beginning of December 1913 while the movements of the landslide body lasted till 1914. L. Sawicki (1917) emphasised very high water saturation of the landslide material which resembled a semi-liquid or liquid substance.

Bystrzyca landslide formed between 3–4.11.1974 (Fig. 4A). This is subsequent, rocky-weathered material landslide, formed within tectonically complicated bedrock (Thiel 1989). The landslide was found on the western slope of Taborówka ridge, near the marginal zone of the Magura nappe overthrust. The underlying rocks comprised clayey shales intercalated with thin and medium bedded sandstone (Inoceramian beds) and shale-sandstone packs. The landslide occupied the lateral, lowered arm of the Taborówka ridge and comprised almost a whole slope from the summit part to the valley floor. The difference in height was 60 m (310 to 370 m a.s.l.), the landslide was 310 m long and ca 140 m wide, and the surface area was 3.6 ha. Two slip zones: at the depth of 3 to 9 m and at 9 to 15 m have been identified in the upper and lower parts of the landslide, respectively. The landslide consists of two parts. The more unstable downslope part, comprising 2/3 of the surface area, is separated from the top part by a distinct 2.5 m high scarp. The head comprises a number of minor, 0.4 m high scarps arranged in an amphitheatre shape. The lower part of the landslide travelled over the distance of ca 7 m and formed a 12 m high escarpment above the valley floor. The landslide damaged a house and a storage building.

This landslide was triggered due to the long-lasting precipitation whose pattern and sums were similar to those recorded in 1913. In 1974 two series of the long-lasting precipitation occurred: since the last week of April until the mid-August and since the last week of September until the beginning of November. At

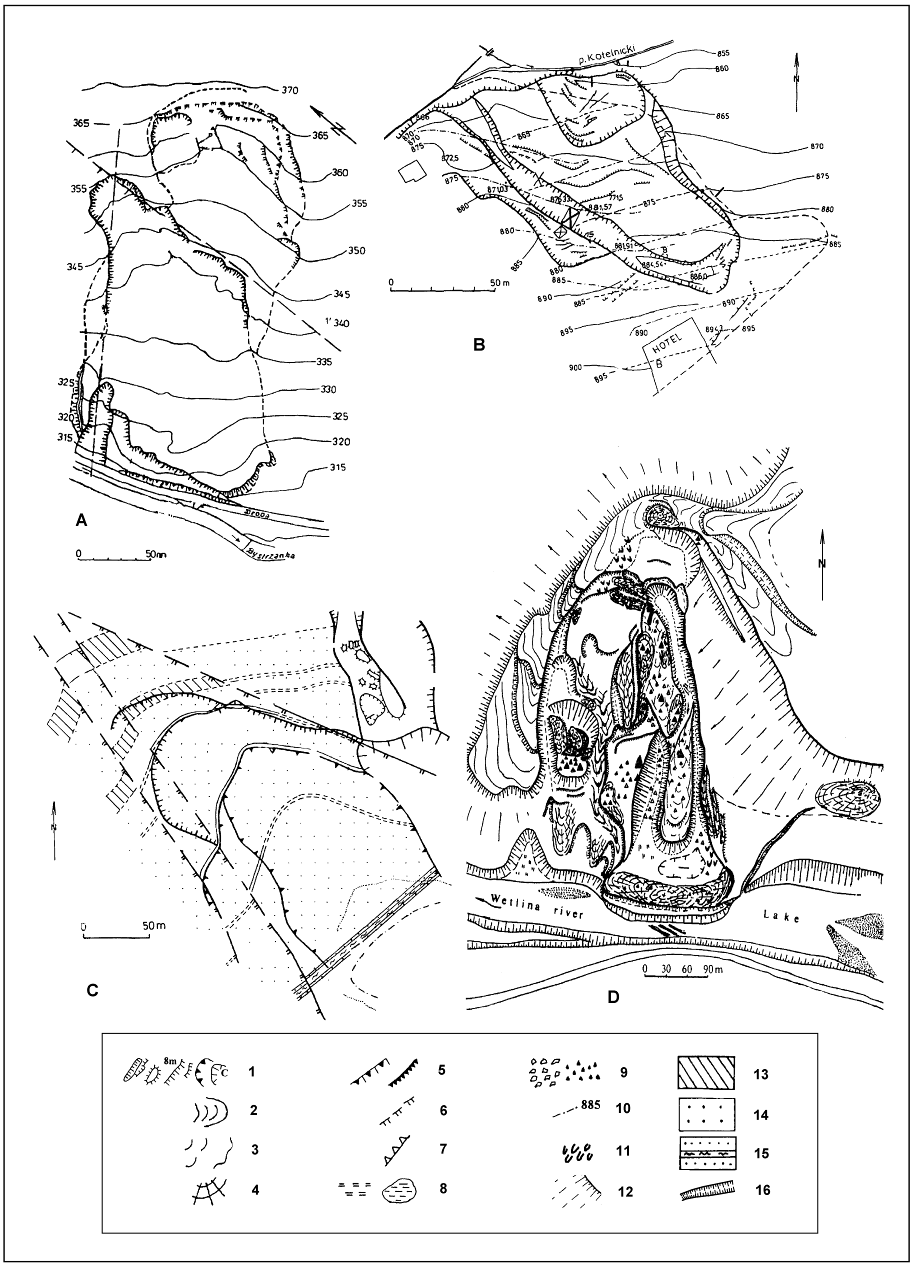

Fig. 4. Sketches of selected Carpathian landslides, A — Bystrzyca (after Thiel 1989), B — Kotelnica (after Bober et al. 1997), C — Lipowica (after Bober et al. 1997), D — Połoma (after Margielewski 1991). 1 — main scarp and head, 2 — landslide front, 3 — landslide humps, 4 — longitudinal bulges, 5 — edge of an artificial scarp, 6 — anticipated shape of the old scarp, 7 — axis of minor fold, 8 — water-logged terrain, 9 — major pile of rock blocks, 10 — isohypses, 11 — creeps, 12 — steep slope, 13 — clayey and marly shale with thin inserts of sandstones, 14 — massive sandstone with inserts of thin mudstone, 15 — thin and massive sandstone with inserts of hard or soft mudstone, 16 — tension cracks

that time intensive mass movements were registered on many slopes in the Carpathians (Bober et al. 1977; Gil and Starkel 1979).

The spring-summer precipitation of the rainy season of 1974 (since 27 May until 13 June), whose cumulated sum was 211 mm, set off a whole series of mass movements with various intensities. At the last day of this precipitation period, i.e. on 13 June, the landslide of Szymbark Folwark was formed. The latter, whose surface area was 0.5 ha, comprised a loamy-debris weathered mantle and partially a shale bedrock, damaged two houses, a road and a regional gas-line.

The landslide of Bystrzyca was formed after rainfalls which lasted since 21 September until the beginning of November (Fig. 5). The landsliding processes started in the mid-October, when, after 3-days of rain of the order of 60 mm, the cumulated precipitation summed to 199 mm. The initiated process of mass movement was interrupted as the precipitation ceased and, then the process became active again during the next more significant precipitation which occurred at the end of October and the beginning of November. The major movements were recorded on 3 and 4 November, when the cumulated precipitation total (since 21 September) was 285 mm (Gil 1994). At the end of October and the beginning of November temperature dropped to almost 0°C and rainy-snow precipitation occurred. This caused an increase in water viscosity and, thus, an outflow form the slope was slow which might also contributed to slope instability.

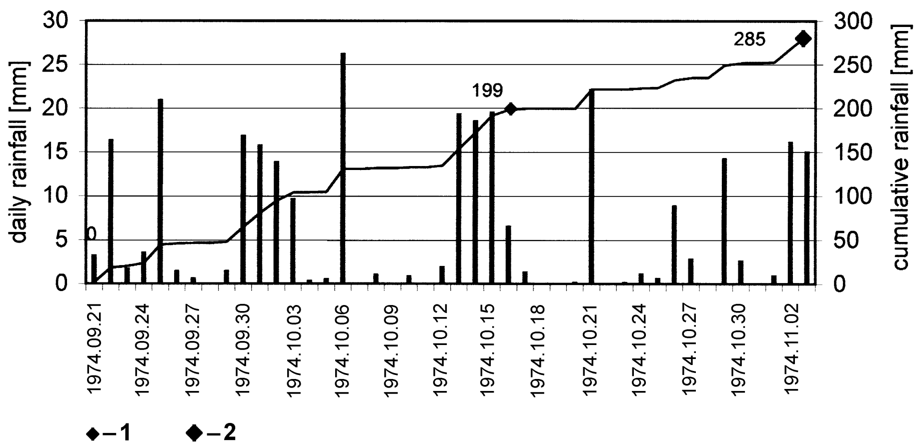

Fig. 5. Precipitation preceding the formation of Szymbark Bystrzyca landslide, 16.09.1974, 3.11.1974 — IGSO PAS Szymbark. 1 — beginning of landsliding, 2 — main phase of landsliding

Landslide of Kotelnica, which was found at 855 to 890 m a.s.l., at the eastern slope of Gubałówka ridge (Inner Carpathians), formed on 18 July 1970 (Fig. 4B). The slopes of Kotelnica were composed of thin-bedded clayey shales and mudstones intercalated with thin sandstones dipping according to the slope inclination (Bober 1975; Thiel 1976). The movement of the rock masses was of structural-consequent type while the transfer of the flysch block occurred over the clay inserted between shale-marly beds. The thickness of this block was 5 m, its length — 135 m, width — 80 m and the surface 1 ha. Landslide material comprises the flysch layers and the loamy cover with rock debris. The landslide main scarp was 5 m high at maximum while the height of the front reached up to 10 m.

The conditions favouring landsliding on the slope of Kotelnica resulted from very high precipitation which occurred in June and July of 1970. The monthly sums at the gauging station at Gubałówka (distant about 1.5 km) were 323.4 mm and 391.2 mm, respectively. This precipitation had a very high daily intensity (Fig. 6). The diurnal precipitation on the day when the landslide was formed reached 149.4 mm. The cumulated precipitation since 14 to18 July was 243 mm (while that calculated since 28 June — 423 mm). Due to ground saturation as well as the slope undercutting by a stream, a rapid movement of earth-rock masses was triggered.

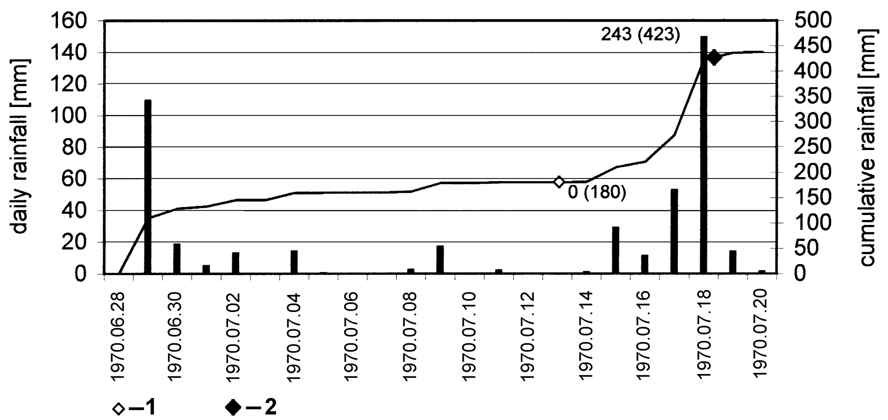

Fig. 6. Precipitation preceding the formation of Kotelnica landslide, 18.07.1970 — Gubałówka. 1 — onset of preceding rainfall, 2 — beginning of landsliding. Sums of precipitation during the movements are given in brackets

Landslide of Lipowica near Dukla was found in the Beskid Niski Mts. and occurred within the Dukla unit (Fig. 4C). The valley-sides of the Jasiołka river are developed of thick and medium bedded sandstones of Cergowa which were intercalated with clayey shales (Gerlach et al. 1958; Thiel 1976). These beds dipped according to the slope inclination. The landside was formed on the slope with the gradient from 15 to 35°, at the heights from 360 to 450 m a.s.l., in the location where the old landslide was formed earlier. In the downslope part there was a quarry and the excavation works carried out there significantly contributed to landsliding. The landslide movements were recorded twice: on 13 May 1957 (Gerlach et al. 1958) and on 18 July 1970 (Thiel 1976). The landslide area was ca 1 ha, its length 270 m, width 70–80 m and the thickness of the slid material was estimated for 25 m at least (G e r l a c h et al. 1958). This was consequent-structural landslide involved rocky-weathered material. The slip surface was thin and comprised shale layer found between the thick-bedded sandstones. The landslide front travelled over the distance of ca 60 m. K. T h i e l (1976) determined the slip surface depth at 10 m in 1970.

The causes of the mass movements in question cannot be determined unambiguously, especially for 1970, when the precipitation sums did not correspond with a possibility of landsliding. On the other hand, the precipitation of 7 April–15 June, which preceded the mass movements, was 163.3 mm. This value might be considered as too low to trigger landsliding (Fig. 7).

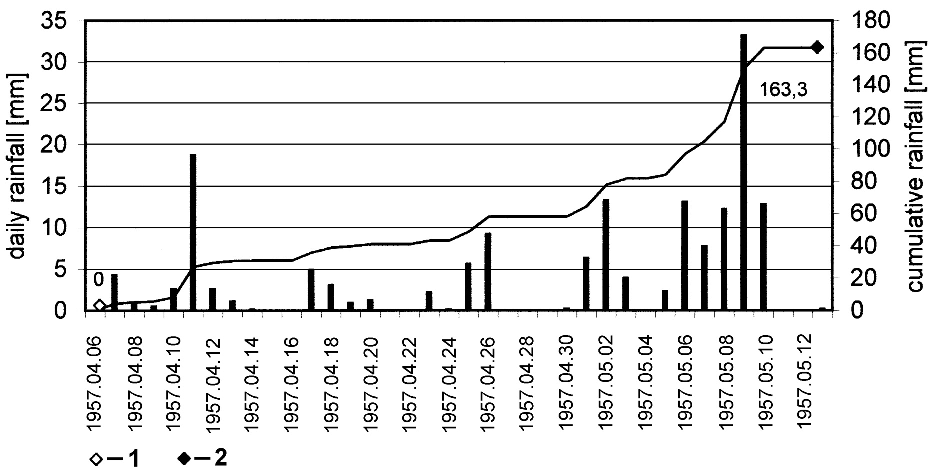

Fig. 7. Precipitation preceding the formation of Lipowica landslide, 13.05.1957 — Iwonicz Zdrój. 1 — onset of preceding rainfall, 2 — beginning of landsliding

Therefore, T. G e r l a c h et al. (1958) believed that the impulse which initiated the movement was a seismic quake registered in the south of Europe on 18 May 1957 as well as disturbances related to the excavation works in the quarry at the foot of the slope. When analysing the height of the precipitation impulse the attention should be paid to snow falls which occurred from 23 April till 15 May and reached 122 mm. The snow blown from the adjacent terrain could have accumulated on the landslide body and formed a thicker cover than it results form the recorded values. This is reasonable especially in the context that the discussed terrain is found to be subjected to very strong winds. Thus, the 3-day lag in the movements when related to the precipitation period can result form a gradual melting of a snow cover. This is a problem in all cases when the relation: precipitation sum — precipitation period — mass movement is considered in winter seasons, especially when the snow cover is long-lasting and variable.

The landslide on Połoma (Fig. 4D) slope in the Wetlinka valley (Western Bieszczady Mts.) was formed on 25–26 July 1980 (Dziuban 1983). An old landslide body was found on this slope. This old landslide was rejuvenated in 1980 and the new movement comprised ca 40% of the old form (Margielewski 1991). The slope of Połoma was composed of the Otryt sandstones of the Silesian nappe with thin intercalations of shales dipping steeply (60°) towards the Wetlina valley. The new landslide was formed within the old colluvium in which the inclination of the rock packs was similar to that of the slope. The discussed landslide is a structural form, comprising rocky-weathered material, where rotational-flow movement occurred. The landslide was found at 639 m a.s.l. and its front in the Wetlinka valley was at 500 m a.s.l. The scarp height reached 10 to 25 m, and the landslide surface was 5.4 ha. The transfer of the landslide masses was over 180 m. This landsliding process caused the river channel to be filled up and a dammed lake to be formed.

The landslide on the slope of Połoma was formed after the long period of high precipitation (Fig. 8). The rain gauging station in Terka (located ca 4 km north of the landslide) registered 263.2 mm in June and 430.1 mm in July. The first series of precipitation of the sum of 160 mm occurred at the end of May and beginning of June while the second series, of the sum of 254 mm, occurred since mid-June till mid-July, and the third precipitation series, which was decisive of landsliding, taking place since 21 July to 26 July summed to 289 mm. Finally, the last day precipitation was 93 mm. All together, in the period from 25.05.1980 till 26.07.1980, precipitation amounted to 719 mm. A mid-June pause in precipitation has been assumed the onset of rainfall which had an influence on the landslide formation. The sum of rainfall of the period from 15 June till 26 July was 553 mm with the most decisive precipitation of the last six days.

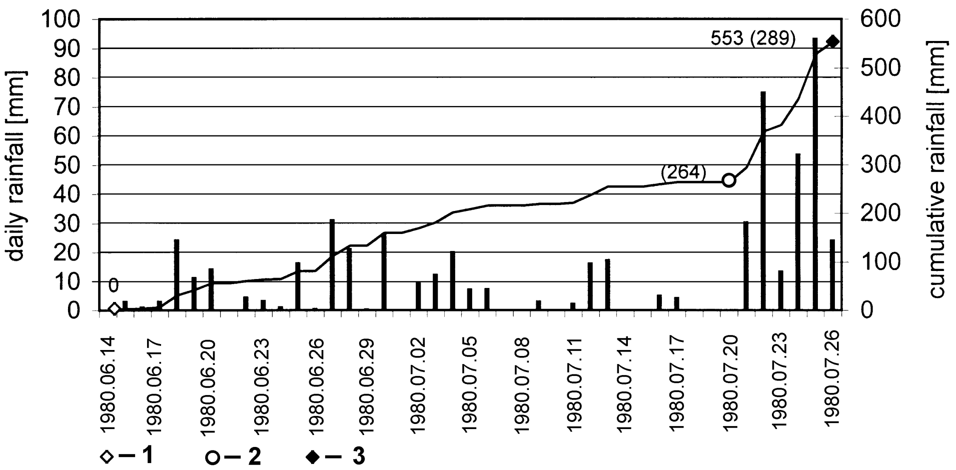

Fig. 8. Precipitation preceding the formation of Połoma landslide, 15–26.07.1980 — Terka. 1 — onset of preceding rainfall, 2 — beginning of the last period of precipitation, 3 — beginning of landslide movement. Sums of precipitation for particular time intervals are given in brackets

The area of the Beskid Żywiecki Mts. is found within the range of the Magura nappe composed of resistant sandstones and less resistant sandstone-shale belonging to the Sub-Magura, Hieroglyphic and Beloveza beds. The inverse relief causes the Beskidy summit parts to comprise the most resistant sandstones (Ziętara 1968). The heights of the main ranges of the Beskidy Mts. exceed 1,000 m a.s.l. with the culmination of Pilsko (1,557 m a.s.l.). The region is dissected by deep valleys with steep valley-sides. In 1958 and 1960 the slopes of the Beskidy Mts. were significantly remodelled by landsliding. Usually, the areas of the old landslide forms were rejuvenated and enlarged. These rock-weathered material landslides are diversified as to a mode of movement, and are found on the slopes with gradients exceeding 15° (Ziętara 1968).

The rainfalls which caused the mass movements in July of 1960 were very high (Fig. 9) in the Beskid Żywiecki Mts. Their first phase lasted from the first days of July till mid-July and amounted to over 200 mm. The second precipitation phase occurred towards the end of July and summed up to 290 mm. All together, the precipitation total of these 25 days reached 523 mm. The Beskid Żywiecki Mts. belong to the regions where precipitation reaches the highest values in the Carpathians (Cebulak 1992) and the daily values exceed 200 mm. The widespread landsliding observed in 1960 was related to extreme rainfalls which occurred in a relative short time span, especially in a few days preceding the mass movement.

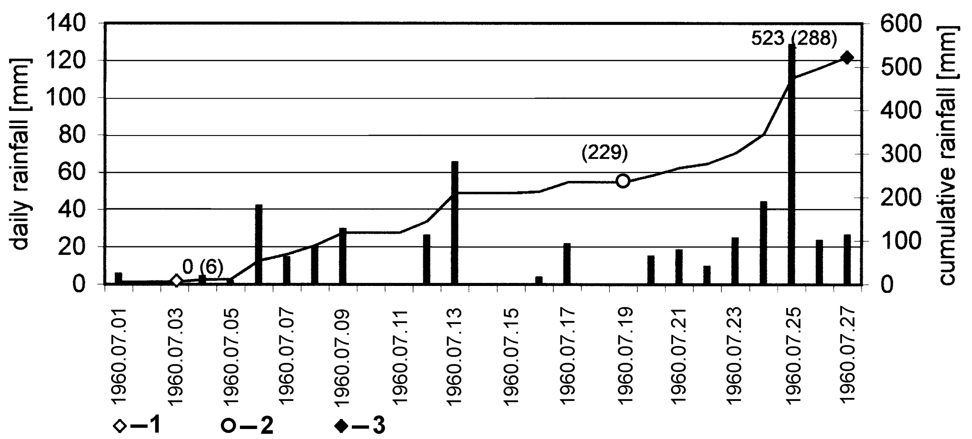

Fig. 9. Precipitation preceding the formation of landslides in Beskid Żywiecki Mts. (Pilsko, Leskowiec, Hala Lipowska) — July 1960. 1 — onset of preceding rainfall, 2 — beginning of the last period of precipitation, 3 — beginning of landslide movement. Sums of precipitation for particular time intervals are given in brackets

THRESHOLD VALUES OF PRECIPITATION PRECEDING MASS MOVEMENTS

Mass movements in the Flysch Carpathians are observed during the extreme precipitation. Determination of a close relationship between the movements and precipitation is an important and challenging task. As results from the examples discussed above, the so called “threshold values” of precipitation initiating mass movements differ significantly (Table 1). The differences result from precipitation sums, rainfall duration and intensity as well as from water circulation conditions in the ground which are conditioned by geological setting (lithology and tectonics), slope gradient, vegetation cover and slope instability due to erosional undercutting or human impact.

In this paper the “precipitation threshold value” is understood as the minimum sum of the rainfall in a given period due to which a mass movement is triggered. Such periods comprise from a few to several dozen days when there are no pauses in precipitation which might cause a groundwater level to fall down.

Table 1

Precipitation thresholds for selected landslides in various regions of the Flysch Carpathians

Landslide	Location/ geologic unit	Geological conditions	Rainfall in the last days preceding a movement onset	Threshold value and duration of precipitation preceding landsliding
Maślana Góra 1.07.1913	Beskid Niski Mts./Magura Nappe	Magura sandstones, variegated shales, Inoceramian beds, Ciężkowice sandstones	64 mm/3 days	219 mm/47 days 318 mm/46 days 537 mm/93 days
Lipowica 13.05.1957	Beskid Niski Mts./Dukla unit	Massive sandstones with thin intercalations of mudstones	81 mm/6 days	63 mm/38 days
Landslides on Pilsko Mts.	Beskid Żywiecki Mts./Magura Nappe	Magura sandstones, sandstones of Sub-Magura, Hieroglyphic and Beloveza beds	289 mm/8 days	523 mm/25 days
Kotelnica 18.07.1970	Podhale/Podhale Flysch	Shales, mudstones, thin sandstones	149 mm/1 day	243 mm/5 days
Szymbark Folwark 13.06.1974	Beskid Niski Mts./Magura Nappe	Inoceramian beds; sandstones, shales	109 mm/4 days	10 mm/16 days
Szymbark Bystrzyca 3.11.1974	Beskid Niski/Magura Nappe	Sandstones/shales of Inoceramian beds; variegated clay-shales	13 mm/2 days	199 mm/26 days 285 mm/45 days
Połoma 26.07.1980	Bieszczady Mts./ Silesian Nappe	Otryt sandstones, clayey shale inserts	289 mm/6 days	537 mm/41 days
Lachowice 27. 07. 2001	Beskid Żywiecki Mts./Magura Nappe	Sub-Magura sandstones and shales	62 mm/1 day	220 mm/7 days
Bystra Podlesie 30.07.2004	Beskid Niski Mts./Magura Nappe	Sandstones and shales of Inoceramian beds	180 mm/4 days	220 mm/10 days

A mutual relation between rainfall and water circulation in the ground (Thiel 1989) is significant for conditioning the mass movements. The long-lasting precipitation with a low intensities which does not exceed the ground infiltration capacity allow for a deep penetration of water into the ground, and this way for deterioration of physical and mechanical parameters of soils, which as a consequence leads to slope instability. On the other hand, during high intensity heavy downpours, significant amounts of water discharge as an overlandflow or throughflow without infiltrating deeper into the ground. Simultaneously, one of necessary conditions of slope instability is a full saturation of the ground with water. According to J. Słupik (1973) this condition for the Carpathians is satisfied when the value of 160 mm is reached (provided the average initial moisture).

The influence of rainfalls and water circulation in the ground on the mass movements was best analyses at Bystrzyca landslide near Szymbark (Gil and Starkel 1979; Thiel 1989; Gil 1997).

During the rainfalls, water infiltrated into the ground, discharged as the overlandflow and throughflow, and also evaporated. When the precipitation totals exceeded 100 to 150 mm, the groundwater level was relatively shallow. The area between the curves 1 and 4 indicates the amount of water left in the ground. During the following increased precipitation, the groundwater level coincided with the ground surface (Fig. 10). As clear from the graph, when the precipitation exceeded 100 mm the amount of water in the ground was almost constant although the oscillations in the groundwater level corresponded well with changes in the precipitation sums. Similar conditions were also observed when the landslide in Szymbark Folwark was formed (Gil 1994, 1997). This constant value demonstrating equilibrium between a water supply and infiltration capacity of the ground is an important input for indicating stability of the slopes under various environmental conditions.

The intensive mass movements were observed after double or triple rises in groundwater level which evidenced that the full water saturation of the ground is necessary to trigger landsliding. The process of manifold saturation of the ground decreases the slope stability in a significant way and, as a consequence, triggers landsliding (Gil and Starkel 1979; Zabuski et al. 2003, 2004).

Based on the studies on the water circulation on the slopes in Szymbark it is evident that conditions necessary for landslide formation in this region are fulfilled by the low intensity precipitation with the sums exceeding 250 mm, duration exceeding 35% of a rainy period and simultaneous recurrent full saturation of the ground (Gil 1997).

In the case of landslides permanently active in summer seasons, although the cumulated sum of precipitation and a number of days with precipitation are similar as in the case of long-lasting rainfalls, only an acceleration of movement or rejuvenation of fragments of partially stabilized landslides was observed. The reasons behind it are high intensity rainfalls which exceed the infiltration capacity of the ground. Thus, water flows out from the slope as rapid overland flow, and

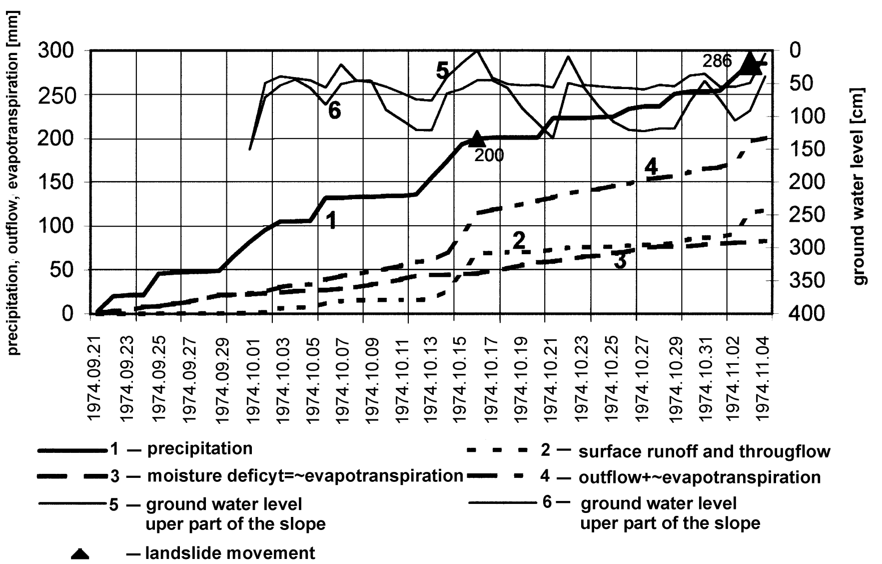

Fig. 10. Water circulation on slopes during movements of Bystrzyca landslide in Szymbark — 21.09.1974–3.11.1974 (after Gil 1994)

throughflow (Słupik 1973; Gil 1997, 1999) and evaporates. In summer season evapotranspiration can amount to over a dozen millimetres of daily water output from the slope. Infiltration affects only the uppermost layer of the ground. Under such circumstances shallow, weathered material landslides, slumps at erosional and denudation scarps as well as debris-mud flows are formed (Gorczyca 2004; Starkel 1960, 1996). Such phenomena are observed during a few hour long precipitation exceeding 100 to 150 mm.

A "threshold precipitation" is also observed when a rainy period lasts a dozen days or more and which terminates with very high precipitation in the final phase. This is well illustrated by Połoma landslide (Fig. 8). The mass movement processes are then of extreme nature.

The heights of precipitation, both preceding and occurring at the moment of landslide formation are important factors causing landsliding. As results from the compiled Table 1, both precipitation heights and its duration are significantly diversified in the Carpathians. W. Rączkowski and T. Mrozek (2002) after compiling about a 100 summer observations of maximum daily, monthly and seasonal precipitation have suggested three precipitation thresholds. If daily precipitation sum is 250 mm, then shallow landslide and debris-mud flows are formed. Due to long-lasting precipitation whose sum exceeds 400 mm many structural landslides are formed and landsliding processes are widespread. Precipitation exceeding 600 mm per month indicates the threshold of extreme landsliding processes. The authors state that deep structural landslides in the Flysch Carpathians are formed when the ground is fully saturated and rainfalls are high. In the Carpathian region

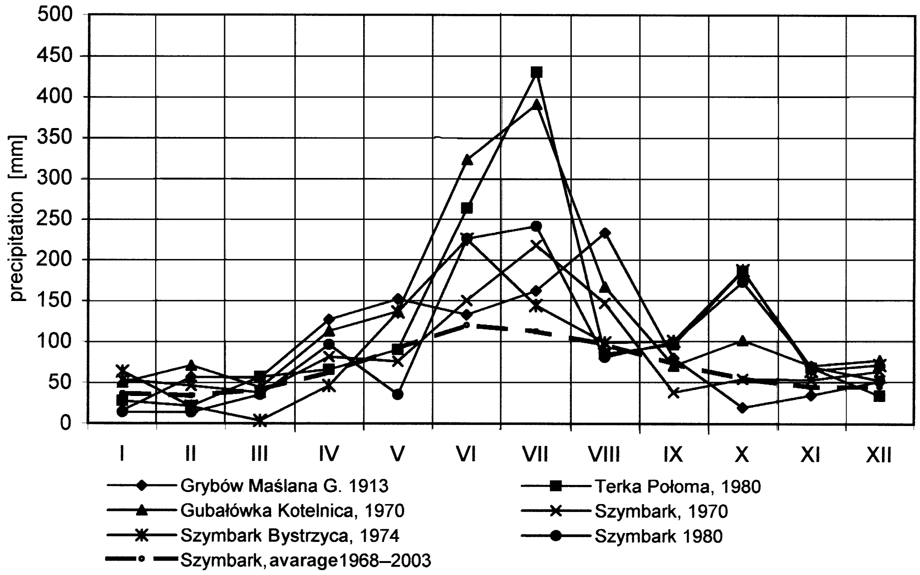

Fig. 11. Monthly precipitation totals in years when the discussed landsliding was observed at the background of precipitation in Szymbark

such situation took place in summer season (April–October) which comprises 65 to 70% of the annual precipitation total (Fig. 11).

In the so called "landslide years" the monthly precipitation totals are 2–3 times higher than the averages of a multi-year period. A specific situation appears when high precipitation cumulates in subsequent months (Maślana Góra 1913, Kotelnica 1970, Szymbark 1974, Połoma 1980). In winter–spring season the mass movements are controlled by a presence of a thick snow cover and its rapid melting combined with rainfalls provided unfrozen bedrock which affects infiltration ability. Such conditions can be exemplified by the observations of 2001 (Rączkowski and Mrozek 2002).

The dependence of landsliding processes on precipitation heights and duration (expressed in days) is presented for selected landslides for which the dates of their formation are known (Fig. 12).

In the case of the particular discussed landslide for which the course of the processes was known (e.g. Maślana Góra, Bystrzyca) more than one precipitation period could have been accepted. The characteristics of the periods of triggering precipitation comprised the rainfall height and the number of days with rainfall when the initial and the main movement phases were recorded. Between the movement phases a momentary stabilisation was observed. This relates to complex landslides, comprising the slopes of differentiated geological structure (e.g. Szymbark Szklarka landslide on the slope of Maślana Góra), where particular

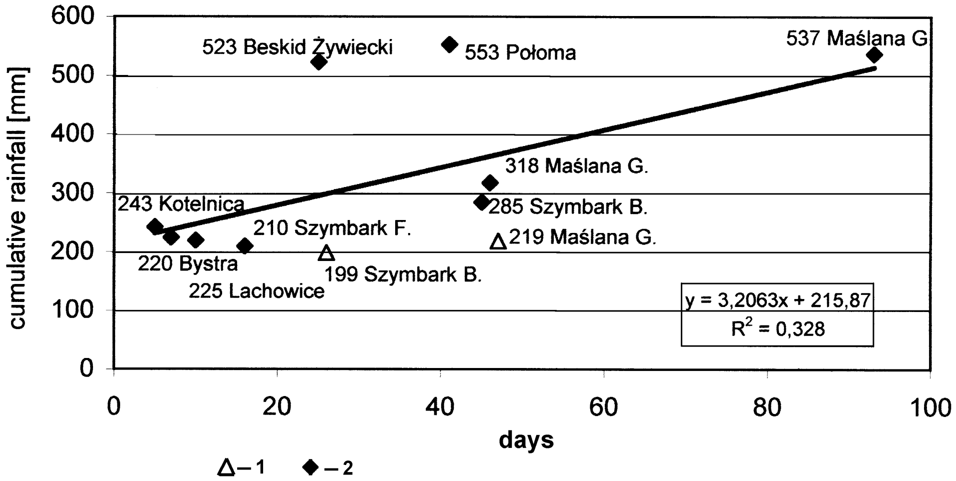

Fig. 12. Dependence of landslides movements on rainfall totals and duration. 1 — beginning of landsliding, 2 — main phase of landsliding

parts of a given landslide, as evidenced by the description of the process, correspond to various precipitation thresholds. The latter cumulate in the catastrophic course of the landsliding process.

Both graph and equation attempt to illustrate the relations which link precipitation and their duration (Fig. 12). Although the number of landslides (objects) presented in the graph is small, certain regularities can be observed.

Landsliding processes usually begin in the Carpathian region after precipitation exceeding 200 mm, lasting a few days provided medium moisture of the bedrock.

There is a considerable difference in the threshold precipitation which triggers landslides on the Beskidy Mts. ridges (where more resistant sandstones predominate) and the landslides occurring in the regions composed of shale-sandstone and shale layers.

Deep-seated rock-weathered material landslides on the Beskidy Mts. ridges composed of sandstones are observed during a longer-lasting precipitation being very high on the last days of rainy period, with the sums exceeding 500 mm in the accepted precipitation series.

Similar structural rock-weathered landslides are formed on the shale-sandstone slopes during precipitation of 200 to 300 mm and lasting form a few to some tens of days.

The differences in the duration and heights of precipitation indicating "threshold values" for landsliding processes result from water circulation pattern and ground saturation in the discussed above areas of the Carpathians.

FINAL REMARKS

Landslide processes belong to main factors modelling the slopes of the Flysch Carpathians. Their common occurrence has been documented in many studies performed by different research centres (Polish Geological Institute, Institute of Geography and Spatial Management Polish Academy of Sciences, Pedagogical University in Cracow), among others in the Landslide Register (PGI Carpathian Branch) demonstrating majority of the Carpathian landslides.

Activity of landslide processes in the Carpathian region is controlled by a very high precipitation with various durations provided a full or partial water saturation of the ground. These conditions affect the course of landsliding which comprises not only the weathered covers but also the bedrock.

Due to geology of the substratum and related relief energy, precipitation initiating the mass movements in the Carpathians is diversified. Determination of the sum of precipitation, satisfying the conditions of full water saturation of the ground or of the weathered cover only, is difficult due to diversified conditions of water circulation in the substratum and due to variability of the sums and intensities of rainfalls in the precipitation period.

Precipitation which lasts one or a few days and reaches 200 to 250 mm, depending on the initial moisture conditions, causes the shallow, weathered material landslides to form on sandstone slope or in the case of the partially stabilised, deeper structural landslide to rejuvenate in the regions composed of shale and shale-sandstone layers.

The cumulated rainfalls, exceeding 250 to 300 mm in a dozen to several tens days, are the reason of formation of structural landslides in the regions where the shales predominate in the substratum. The structural landslides might comprise the old, stabilised forms and new ones.

The precipitation necessary for triggering the landslide on the Beskidy slopes, where sandstones predominate, is 500 mm. The rainfalls of this order of magnitude set off catastrophic mass movements.

The landslide processes comprising the larger regions or these landsliding processes which are rapid belong to extreme events whose results are catastrophic especially for the management in the given region.

The accepted “precipitation threshold values” should be substantiated by other data reflecting differentiation of the Carpathian environment and by consideration of recurrence of the processes. This will be important for a perspective management strategy, forecasting of natural hazard and planned mitigation.

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REFERENCES

Bober L., 1975. *Metody określania warunków hydrogeologicznych na zboczach osuwiskowych w Lipowicy, Kotelnicy, i Dobczycach*. Materiały Badawcze, Seria Specjalna 4, Instytut Meteorologii i Gospodarki Wodnej, Warszawa.

Bober L., Chowaniec J., Oszczypko N., Wójcik A., Witek K., 1977. *Geologiczne warunki rozwoju osuwiska w Brzeżance koło Strzyżowa*. Przegląd Geologiczny 7, 372–377.

Bober L., Thiel K., Zabuski L., 1997. *Zjawiska osuwiskowe w polskich Karpatach fliszowych*. Instytut Budownictwa Wodnego PAN, Gdańsk–Warszawa–Kraków, 102 pp.

Caine N., 1980. *The rainfall intensity — duration control of shallow landslides and derbis flows*. Geografiska Annaler 62A, 1–2, 23–17.

Cebulak E., 1992. *Maksymalne opady dobowe w dorzeczu górnej Wisły*. Zeszyty Naukowe UJ, Prace Geograficzne 90, 79–96.

Dziuban J., 1983. *Osuwisko Połoma*. Czasopismo Geograficzne 4, 3, 369–376.

Froehlich W., Starkel L., 1987. *Norma land extreme monsoon rains — their role in the shaping of the Darjeeling Himalaya*. Studia Geomorphologica Carpatho-Balcanica 21, 129–160.

Froehlich W., Starkel L., 1991. *Wartości progowe w ewolucji rzeźby Karpat fliszowych i Himalajów Dardżylińskich*. Conference Papers IG i PZ PAN 14, 49–58.

Gerlach T., Pokorny J., Wolnik R., 1958. *Osuwisko w Lipowicy*. Przegląd Geograficzny 4, 30, 685–700.

Giannecchini R., 2005. *Rainfall triggering soil slips in the southern Aquan Alps*. Advances in Geosciences 2, 21–14.

Gil E., 1994. *Meteorologiczne i hydrologiczne warunki ruchów osuwiskowych*. Conference Papers IG i PZ PAN 20, 89–102.

Gil E., 1997. *Meteorological and hydrological conditions of landslides in the Polish Flysch Carpathians*. Studia Geomorphologica Carpatho-Balcanica 31, 143–158.

Gil E., 1999. *Obieg wody i spłukiwanie na fliszowych stokach użytkowanych rolniczo w latach 1980–1990*. Zeszyty Instytutu Geografii i Przestrzennego Zagospodarowania Polskiej Akademii Nauk 60, 78 pp.

Gil E., Kotarba A., 1977. *Model of slide slope evolution in flysch mountains (An example drawn from the Polish Carpathians)*. Catena 4, 3, 233–248.

Gil E., Starkel L., 1979. *Long-term extreme rainfalls and their role in the modelling of flysch slopes*. Studia Geomorphologica Carpatho-Balcanica 13, 207–120.

Glade T., 1998. *Establishing the Frequency and Magnitude of Landslides — triggering Rainstorm Events in New Zealand*. Environmental Geology 35 (2/3), 160–174.

Gorczyca E., 2004. *Przekształcenie stoków fliszowych przez procesy masowe podczas katastrofalnych opadów (dorzecze Łososiny)*. Wydawnictwo Uniwersytetu Jagiellońskiego, Kraków, 101 pp.

Govi M., Sorzana P., Tropeano D., 1982, *Landslide mapping as evidence of extreme regional events*. Studia Geomorphologica Carpatho-Balcanica 15, 81–98.

Innes J. L., 1983. *Debris flows*. Progress in Physical Geography 7, 469–501.

Jakubowski K., 1964. *Płytkie osuwiska zwietrzelinowe na Podhalu*. Prace Muzeum Ziemi 6, 113–146.

Jakubowski K., 1965, *Wpływ pokrycia roślinnego oraz opadów atmosferycznych na powstanie osuwisk zwietrzelinowych*. Przegląd Geologiczny 13, 395–398.

Katalog osuwisk — województwo krakowskie i rzeszowskie, 1975, Praca zbiorowa, OK Państwowy Instytut Geologiczny Kraków.

Kleczkowski A., 1955. *Osuwiska i zjawiska pokrewne*. Wydawnictwa Geologiczne, 93 pp.

Margielewski W., 1991, *Land sidle forms on Połoma Mountain in the Sine Wiry Nature Reserve, West Bieszczady*. Ochrona Przyrody 49, I, 23–19.

Mrozek T., Rączkowski W., Limanówka D., 2000. *Recent landslides and triggering climatic conditions in Laskowa and Pleśna regions, Polish Carpathians*. Studia Geomorphologica Carpatho-Balcanica 34, 89–109.

Rączkowski W., Mrozek T., 2002. *Activiting of landsliding in the Polish Flysch Carpathians by the end of the 20th century*. Studia Geomorphologica Carpatho-Balcanica 36, 91–111.

Sawicki L., 1917. *Osuwiska ziemne w Szymabrku i inne zsuwy powstałe w r. 1913 w Galicji zachodniej*. Rozprawy Wydziału Matematyczno-Przyrodniczego Akademii Umiejętności, 3, A, 227–313.

Sikora W., 1970. *Budowa geologiczna płaszczowiny magurskiej między Szymbarkiem Ruskim a Nawojową*. Instytut Geologiczny, Biuletyn 235, Z badań geologicznych w Karpatach, 13, 127 pp.

Słupik J., 1973. *Zróżnicowania spływu powierzchniowego na fliszowych stokach górskich*. Dokumentacja Geograficzna 2, 118 pp.

Starkel L., 1960, *Rozwój rzeźby Karpat fliszowych w holocenie*. Prace Geograficzne Instytutu Geografii Polskiej Akademii Nauk 22, 239 pp.

Starkel L., 1976. *The role of extreme (catastrophic) meteorological events in the contemporaneous evolution of slopes*, [in:] *Geomorphology and Climate*, J. Wiley and Sons, London, 203–146.

Starkel L., 1996. *Geomorphic role of extreme rainfalls in the Polish Carpathians*. Studia Geomorphologica Carpatho-Balcanica 30, 21–38.

Terlin M. T. J., 1997. *Hydrological landslide triggering in ash-covered slopes of Manizales (Colombia)*. Geomorphology 20, 165–175.

Thiel K., 1976. *Badanie i prognozowanie stateczności zboczy skalnych*. Prace Instytutu Budownictwa Wodnego PAN 2, 83 pp.

Thiel K., (ed.) 1989. *Kształtowanie fliszowych stoków karpackich przez ruchy masowe*. Prace Instytutu Budownictwa Wodnego PAN, Gdańsk–Kraków, 91 pp.

Zabuski L., Thiel K., Bober L., 1999. *Osuwiska we fliszu Karpat polskich*. Wydawnictwo Instytutu Budownictwa Wodnego PAN, Gdańsk, 171 pp.

Zabuski L., Gil E., Rączkowski W., Wójcik A., 2003. *Badania reprezentatywnego procesu osuwiskowego w masywie fliszu karpackiego — osuwisko eksperymentalne w Beskidzie Niskim*. Sprawozdanie z Projektu Badawczego KBN 8 T12B 047 20, Archiwum Instytutu Budownictwa Wodnego Polskiej Akademii Nauk, Gdańsk, 35 pp.

Zabuski L., Gil E., Bochenek W., 2004, *Interdependence between groundwater level and displacement of the landslide slope*. Polish Geological Institute, Special Issue 15, 39–42.

Ziętara T., 1968. *Rola gwałtownych ulew i powodzi w modelowaniu rzeźby Beskidów*. Prace Geograficzne Instytutu Geografii Polskiej Akademii Nauk 60, 116 pp.

STRESZCZENIE

Eugeniusz Gil, Michał Długosz

WARTOŚCI PROGOWE OPADÓW URUCHAMIAJĄCYCH WYBRANE GŁĘBOKIE OSUWISKA W KARPATACH FLISZOWYCH

Procesy osuwiskowe należą do najważniejszych procesów morfogenetycznych modelujących środowisko przyrodnicze Karpatach fliszowych, obejmując około 20% stoków. Warunkiem sprzyjającym procesom osuwiskowym jest budowa geologiczna Karpat, w postaci na przemian ułożonych warstw piaskowcowych i łupkowych, pofałdowanych i pociętych uskokami. Istotnym czynnikiem jest również rzeźba terenu, wyrażona wysokościami względnymi, nachyleniem i długością stoków oraz ich relacją do biegów i upadów warstw skalnych.

Głównymi dynamicznymi czynnikami inicjującymi ruchy osuwiskowe są opady deszczu lub topnienie śniegu oraz obieg wody w podłożu. Warunkiem powstania ruchów osuwiskowych jest pełna saturacja podłoża (zwierciadło wód gruntowych kształtuje się na poziomie lub w pobliżu powierzchni terenu), która dla pokryw zwietrzelinowych w Karpatach określana jest na około 160 mm. Stan pełnej saturacji występuje najczęściej w okresie długotrwałych, ciągłych (lub z małymi przerwami) opadów, których natężenie umożliwia głęboką infiltrację wody w podłoże (ryc. 10). Podczas gwałtownych, wysokich opadów przeważa spływ powierzchniowy i płytka infiltracja, będąca na ogół przyczyną płytkich zsuwów pokrywy zwietrzelinowej. Duży wpływ na stan nasycenia wodą podłoża w okresie letnim ma ewapotranspiracja, odprowadzająca znaczne ilości wody.

Wysokość opadów inicjujących ruchy osuwiskowe określana jest jako „próg opadowy" (Froehlich, Starkel 1987). Dla regionów o różnej budowie geologicznej wysokość opadów inicjujących ruchy masowe ma różną wartość (Govi et al. 1982; Rączkowski, Mrozek 2002). Natomiast od natężenia i czasu trwania opadów zależy mechanizm i typ ruchu osuwiskowego.

Wysokość progów opadowych inicjujących powstanie osuwisk przedstawiono na przykładzie kilku osuwisk, położonych w różnych regionach fizyczno-geograficznych Karpat: Beskid Niski, Beskid Żywiecki, Podhale, Bieszczady. Dla tych osuwisk, o znanych datach powstania, wyznaczono okresy opadowe — sumę i czas trwania opadów — które były przyczyną ich ruchu (ryc. 3, 5, 6, 7, 8, 9).

Wartości progowe opadów inicjujących procesy osuwiskowe są różne w poszczególnych regionach fizyczno-geograficznych Karpat (ryc. 12).

Procesy osuwiskowe, w warunkach średniej wilgotności podłoża, zaczynają się w obszarze karpackim po opadach o wysokości przekraczającej 200 mm i czasie trwania od kilku do kilkunastu dni.

Na stokach zbudowanych z warstw łupkowo-piaskowcowych, strukturalne, skalno-zwietrzelinowe osuwiska, powstają po opadach o wysokość 250–300 mm, trwających od kilku do kilkudziesięciu dni.

Na stokach grzbietów beskidzkich z przewagą warstw piaskowcowych, głębokie, skalno-zwietrzelinowe osuwiska powstają, kiedy skumulowana suma opadów przekracza 500 mm w ciągu kilkudziesięciu dni, a maksimum opadów występuje w ostatnich dniach okresu deszczowego.

Ruchy osuwiskowe mają charakter procesów o charakterze ekstremalnym i związane są najczęściej z opadami okresu letniego, w którym sumy miesięczne przekraczają 2–3-krotnie średnie wartości.

Współcześnie obserwowane ruchy osuwiskowe najczęściej odnawiają stare, ustabilizowane osuwiska, powiększając je o przylegające do nich części stoków stabilnych.

Powszechność procesów osuwiskowych w Karpatach stanowi poważny i bardzo trudny do rozwiązania problem gospodarczy w regionie.